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# Evaporation from rain-wetted forest in relation to canopy wetness, canopy cover, and net radiation

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**Abstract.** Evaporation from wet canopies is commonly calculated using  $E_{PM}$ , the Penman-Monteith equation with zero surface resistance. However, several observations show a lower evaporation from rain-wetted forest. Possible causes for the difference between  $E_{PM}$  and experiments are evaluated to provide rules for the simulation of rainfall interception by forest canopies. The evaluation is executed using a micrometeorological model with a detailed representation of the forest canopy. Simulated results are compared with experimental results. In spite of theoretical reservations the evaporation of completely wet forest appears to agree with  $E_{PM}$ . Evaporation from wet forest appears mainly dependent on net radiation. Rainfall interception is related to evaporation from the canopy. Evaporation from the canopy appears proportional with the square root of canopy cover and sensitive to canopy wetness. An accurate estimate of canopy wetness is needed to use  $E_{PM}$  for the calculation of evaporation from rain-wetted forest.

## 1. Introduction

Evaporation from rain-wetted forests affects the water use, as a significant fraction of rainfall on forest is intercepted by the canopy and evaporated without reaching the soil (see Figure 1). According to *Hutjes et al.* [1990] and *Loustau et al.* [1992] the main factors influencing the amount of rainfall interception are storage capacity of the canopy and evaporation rate. A simple and accurate method is desired to calculate evaporation from rain-wetted canopies. Commonly, evaporation from wet canopies is calculated with  $E_{PM}$ , the Penman-Monteith estimate of evaporation with zero surface resistance. The Penman-Monteith equation has successfully been applied to calculate transpiration, but the results of  $E_{PM}$  for the evaporation from rain-wetted forest are less convincing. Recent experiments by *Lankreijer et al.* [1993] and *Valente et al.* [1997] showed that the interception of rainfall by forest was overestimated when  $E_{PM}$  was used.

Estimations of evaporation based on rainfall interception experiments are indirect as other parameters affect the results as well. Direct atmospheric measurements of evaporation from wet surfaces during rain are preferred, but these measurements are exceptional owing to the difficulty in maintaining dry instrumentation. *Stewart* [1977] used the Bowen ratio method above pine forest during daytime conditions. The average result ( $E = 0.19 \text{ mm h}^{-1}$ ) agreed with  $E_{PM}$  found by *Gash* [1979] for the same forest, suggesting  $E = E_{PM}$ . However, as the measurements were restricted to daytime with the available energy exceeding  $20 \text{ W m}^{-2}$  and  $E_{PM}$  was calculated for the complete data set, the apparent agreement implies  $E < E_{PM}$  [*Klaassen et al.*, 1998]. *Lindroth* [1993] used  $E_{PM}$  to calculate the aerodynamic roughness of short willow forest using evap-

oration measurements when all leaf wetness sensors indicated a wet canopy. The result of a realistic roughness suggests  $E = E_{PM}$ . *Mizutani et al.* [1997] obtained  $E \sim E_{PM}$  with eddy correlation measurements of evaporation from deciduous forest, but the Bowen ratio measurements resulted in lower evaporation by a factor of 2. Eddy correlation measurements by *Gash et al.* [1999] above sparse pine forest and by *Lankreijer et al.* [1999] above boreal pine forest resulted in values below  $E_{PM}$ . Thus most of the experiments indicate  $E < E_{PM}$  for rain-wetted forest, but others still suggest that  $E = E_{PM}$ . Could it be that serious deviations from  $E_{PM}$  occur under rainy conditions, or should we explain the observed deviations by the difficulty to obtain accurate measurements under rainy conditions? Given the contradicting results of experimental investigations, it was decided not to execute an additional field experiment to solve this question. Instead, available observations have been compared to a theoretical evaluation of the accuracy of  $E_{PM}$  for the calculation of evaporation from rain-wetted forest. The accuracy of  $E_{PM}$  is evaluated using an estimation of possible errors resulting from assumptions in the Penman-Monteith approach. The basic assumption in the Penman-Monteith approach is an equal source distribution of surface fluxes. This assumption is tested using a multilayer canopy model as this type of model can simulate differences in the vertical source distribution [*Watanabe and Mizutani*, 1996]. Different source distributions may arise in the horizontal direction when canopy cover is incomplete, so a two-dimensional (2-D) model is preferred. The 2-D model of *Klaassen* [1992] was selected, as this model was already used to simulate interception near a forest edge [*Klaassen et al.*, 1996].

Several corrections on  $E_{PM}$  have been proposed, for instance, a correction for incomplete canopy wetness [*Rutter et al.*, 1971], incomplete canopy cover [*Gash et al.*, 1995], and a correction on the resistance for water vapor transport [*Lankreijer et al.*, 1993]. Finally, it has been assumed that  $E_{PM}$  is not an independent boundary condition but instead is sensitive to actual evaporation [*Bouchet*, 1963]. The performance of these adjustments on  $E_{PM}$  is also evaluated in this study.

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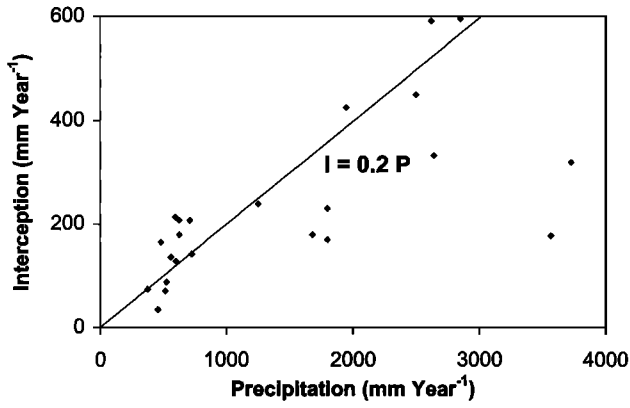


Figure 1. Observed yearly rainfall interception versus precipitation (data from Choudhury and DiGirolamo [1998]). Data well below line  $I = 0.2 P$  were measured at tropical locations.

## 2. Method

### 2.1. Evaporation From a Single Layer Canopy

When all fluxes arise from the same level in the canopy, evaporation from wet vegetation is given by the Penman-Monteith equation with zero surface resistance:

$$\lambda E_{PM} = \frac{\Delta A + \rho C_p (e_a^* - e_a)/r_a}{\Delta + \gamma}, \quad (1)$$

where  $\Delta$  is the slope of the saturated vapor pressure with temperature ( $\text{Pa K}^{-1}$ );  $A$  is available energy;  $\rho$  is the air density ( $\text{kg m}^{-3}$ );  $C_p$  is the heat capacity of air at constant pressure, taken as  $C_p = 1005 \text{ J kg}^{-1}$ ;  $r_a$  is the atmospheric transport resistance ( $\text{s m}^{-1}$ );  $\gamma$  is the psychrometer constant, taken as  $\gamma = 66 \text{ Pa K}^{-1}$ ; and  $e_a$  ( $e_a^*$ ) is the (saturated) vapor pressure (Pa).

Deviations from (1) occur when the canopy is not completely wet or when the canopy cover is not complete. For these situations the following empirical corrections have been derived:

Evaporation from vegetation with partial canopy cover is assumed to be proportional to canopy cover ( $c$ ) as proposed by Gash *et al.* [1995]:

$$E_c = c E_{PM}. \quad (2)$$

The subscript  $c$  is included to emphasize that (2) is restricted to evaporation from the canopy. The understorey component is denoted by  $E_s$ , and the sum of understorey and canopy evaporation is called total evaporation.

Evaporation of partially wet canopies is assumed to be proportional to water storage on the canopy, according to Rutter *et al.* [1971]

$$E_w = (C S^{-1}) E_{PM}, \quad (3)$$

where  $C$  is the actual water storage and  $S$  is the maximum water storage of the canopy (both in mm). The subscript  $w$  means that (3) is restricted to evaporation from the wet part of the canopy. Additional evaporation arises as transpiration from the dry part of the canopy [Larsson, 1981; Bosveld and Bouten, 1999].

The evaporation rate according to  $E_{PM}$  consists of an energy-driven part, related to  $A$ , and a transport-driven part, related to  $(e_a^* - e_a)/r_a$ . The high aerodynamic roughness of

forest might favor dominance of the transport term [Teklehaimanot *et al.*, 1991]. By contrast, the aerodynamic resistance suppresses transport from shielded surfaces. In the limit of an infinite  $r_a$  the transport term vanishes, and (1) simplifies to the equilibrium evaporation  $E_{eq}$ , given by

$$\lambda E_{eq} = \frac{A}{1 + \gamma/\Delta}. \quad (4)$$

The accuracy of (4) will be evaluated for determining evaporation from wet, shielded surfaces. Also, differences from equilibrium evaporation will be used to estimate the significance of atmospheric transport on evaporation.

In (3) the evaporation during incomplete wetness is related to  $E_{PM}$ . Evaporation results in a reduction of humidity deficit in the atmosphere and thus to a decrease of  $E_{PM}$ . The complementary relation

$$E + E_{PM} = 2E_{PO}, \quad (5)$$

where  $E_{PO}$  is the potential evaporation, has been proposed by Bouchet [1963] to estimate the interaction between actual evaporation and  $E_{PM}$ . The complementary relation results in a sensitivity  $\varepsilon = \partial E_{PM}/\partial E = -1$ . By contrast, the assumption of  $E_{PM}$  being an independent boundary condition results in  $\varepsilon = 0$ . The value of  $\varepsilon$  will be estimated using simulations of atmospheric interactions during rainfall.

### 2.2. Evaporation From a Two-Dimensional Canopy

Simulation of the vertical distribution of heat and momentum sources in the canopy enables a check on the assumption of equal resistances for transport of sensible and latent heat and for momentum in the Penman-Monteith equation. The vertical source distributions are simulated by representing the vegetation with several stacked leaf layers with air flowing around the leaves, using the surface layer model of Klaassen [1992]. The wind velocity is calculated from common flux-gradient relations, but a correction by Li *et al.* [1985] is used to enhance wind velocity in the lower part of the forest canopy. The exchange between leaves and the surrounding air is calculated with the Penman-Monteith equation (1). On the leaf scale the atmospheric resistance is caused by a leaf boundary resistance  $r_b$  ( $\text{m s}^{-1}$ ), given by Pearman *et al.* [1972]

$$r_b = 90 \sqrt{l_w/u}, \quad (6)$$

where  $l_w$  is the effective leaf width, generally taken as 0.05 m. Results of (6) are in close agreement with experimental results by Brenner and Jarvis [1995]. The leaf boundary resistance is restricted to heat transport and is not used for momentum transport. The difference between heat and momentum transport has initiated a correction on the atmospheric resistance in the Penman-Monteith equation [Lankreijer *et al.*, 1993, 1999] and motivates the estimation of the sensitivity of evaporation to the leaf boundary resistance in the present study.

Horizontal variability of the forest is included in the model to validate the corrections on  $E_{PM}$  for incomplete wetness and canopy cover (equations (2) and (3)). The heterogeneity is simulated assuming strips of vegetation with different surface characteristics perpendicular to the wind direction. In this way, a two-dimensional surface is simulated with constant properties in the horizontal direction perpendicular to the atmospheric flow and variability in the vertical and in the flow direction. Horizontal variability in surface fluxes arises by variations in surface conditions and by variations in atmospheric

properties like temperature, humidity, and wind velocity. The variability in the atmospheric surface layer is calculated with mixing length theory and common flux-gradient relations. The mixing length is advected with the flow and adjusts slowly to the underlying surface [Klaassen, 1992].

The simulations start with atmospheric profiles that are adjusted to a smooth surface. After the initialization the air is simulated to enter the forest, and the simulation continues until the fluxes in the surface layer converge to the fluxes from the forest. In case of patchy forest the patches are repeated in the wind direction, and the surface layer converges to the spatially averaged surface fluxes. The upper boundary is taken constant in the flow direction, and convergence to a constant boundary implies the assumption of temporal stationarity.

In most situations with a wet surface a stable surface layer develops in which transport is reduced. A problem arises as constant fluxes in the vertical imply an increase of stability with height above the surface. Strong stability during rain at elevated heights is not realistic for the following reasons: (1) the assumption of stationarity is seldom fulfilled during stable situations, owing to gradual cooling of the surface and the lower parts of the surface layer, and (2) the assumption of negligible heat sources in the surface layer does not hold during rain because of heat transfer between falling rain drops and the surrounding air. Extreme stability is prevented by decreasing the upper boundary height from 200 m in the original model to 125 m. Using this thickness of the surface layer, convergence is reached after a fetch of 5 km. For heterogeneous surfaces the available energy  $A$  might vary, but incoming radiation is assumed to be constant. Therefore the available energy is calculated using

$$A = R_n - G = (1 - \alpha)R_s + \varepsilon(R_l - \sigma T_s^4) - G, \quad (7)$$

where  $\alpha$  is the albedo and  $\varepsilon$  is the emissivity of the surface;  $\sigma$  is the Stephan-Boltzmann constant ( $\sigma = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$ );  $R_n$ ,  $R_s$ , and  $R_l$  are the net radiation, incoming short wave and long wave radiation, respectively; and  $G$  is the ground heat flux. All energy fluxes are in  $\text{W m}^{-2}$ . The ground heat flux is assumed to be 10% of the net radiation reaching the soil. Net radiation within the forest is calculated assuming an exponential extinction of net radiation due to absorption by leaves. The extinction coefficient is taken as 0.5 [Jarvis and Leverenz, 1983].

### 2.3. Input Data for Simulations

The standard input data of the 2-D model are given in Table 1. Note that atmospheric humidity is expressed as relative humidity to account for the increase of saturated vapor pressure with temperature. The data are representative of summer daytime conditions in a temperate moist climate during rain above forest. The understory data are used for the forest floor and for the unforested patches and represent a grass-covered soil. As the simulated ratio  $E/E_{\text{PM}}$  is only marginally sensitive to forest height and vertical distribution of the leaves, the presentation is restricted to fixed values of these parameters. This does not mean that evaporation ( $E$ ) is insensitive to the height of forest, rather it means that  $E$  and  $E_{\text{PM}}$  show the same increase with forest height. Unless otherwise stated,  $E_{\text{PM}}$  is calculated at the first atmospheric grid level above the forest at 24 m height to keep in line with common measurement heights. By contrast, the input data of wind velocity, temperature, and humidity are given at the upper boundary of the

**Table 1.** Input Data for Simulations

| Parameter  | Value                 |
|--|-----------------------|
| <b>Meteorological data at 125-m height</b>             |                       |
| Air temperature  | 20°C                  |
| Relative humidity                                      | 90%                   |
| Wind velocity  | 5 m s <sup>-1</sup>   |
| Incoming shortwave radiation                           | 100 W m <sup>-2</sup> |
| Incoming longwave radiation <sup>a</sup>               | 400 W m <sup>-2</sup> |
| <b>Forest data</b>                                     |                       |
| Albedo   | 0.11                  |
| LAI  | 5                     |
| <b>Vertical distribution of LAI at center of layer</b> |                       |
| 2 m  | 0.03                  |
| 4 m  | 0.07                  |
| 6 m  | 0.3                   |
| 8 m  | 0.7                   |
| 10 m   | 1.0                   |
| 13 m   | 1.1                   |
| 16 m   | 1.1                   |
| 20 m   | 0.7                   |
| <b>Understorey data</b>                                |                       |
| Albedo   | 0.16                  |
| Aerodynamic roughness                                  | 0.03 m                |

<sup>a</sup>Equivalent to a sky radiation temperature, 17°C.

model at 125 m height. At this level the influence of the forest on atmospheric conditions is expected to be small, so atmospheric input data can be considered as independent boundary conditions.

The sensitivity of evaporation to partial canopy cover and wetness is simulated to evaluate the single-layer expressions. Multiple canopy representations are used to estimate the sensitivity of evaporation to canopy architecture.

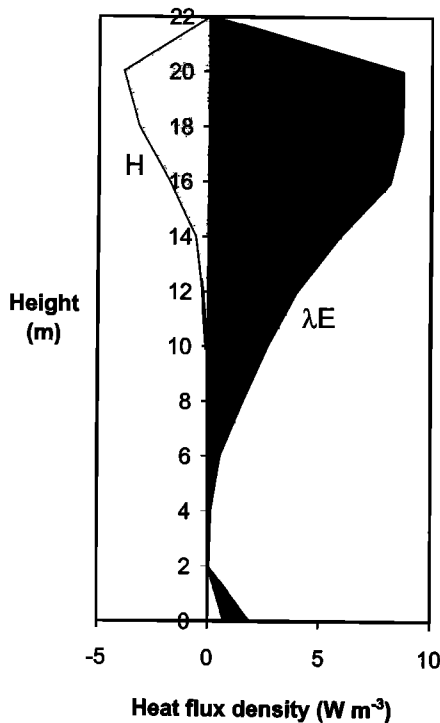
The influence of canopy cover, or forest density, is simulated by two canopy representations. The first representation uses a homogeneous canopy with a random orientation of leaves and random horizontal spacing. Canopy cover  $c$  is then given by

$$c = 1 - \exp(-\text{LAI}/2), \quad (8)$$

where LAI is the leaf area index ( $\text{m}^2 \text{ m}^{-2}$ ).

The second representation simulates patchy vegetation as an alternation of infinite long-forested and nonforested strips at right angles to the wind direction. The forested strips are simulated as homogeneous canopies. The canopy cover of patchy vegetation is taken equal to the forested fraction. The combined length of a forested and a nonforested strip of patchy forest is arbitrarily set at 50 m in the direction of the wind. Klaassen [1992] showed that aggregated fluxes are almost independent for this length scale if the length scale is several tens of meters or smaller.

Canopy wetness is defined as the ratio between wet leaf area and total leaf area. Assuming a constant thickness of the water layer, canopy wetness equals the ratio  $C/S$  between actual and maximum water storage. Partial canopy wetness is simulated in three ways to estimate the influence of the source area distribution on evaporation. With the first method, denoted by "wetting," a fully wet upper part of the canopy is simulated above a dry bottom part. Variations in canopy wetness result from variations in the height of the boundary between wet and dry. The second method, denoted by "drying," simulates a dry upper part above a wet lower part. The methods are denoted by wetting and drying, as these processes tend to start at the upper part of the canopy. In reality, the separation between wet and



**Figure 2.** Simulated vertical source density distribution of sensible and latent heat in a rain-wetted homogeneous forest. Density is the source strength divided by the vertical height interval. The ground surface fluxes are presented as density over the lowest 2 m height interval.

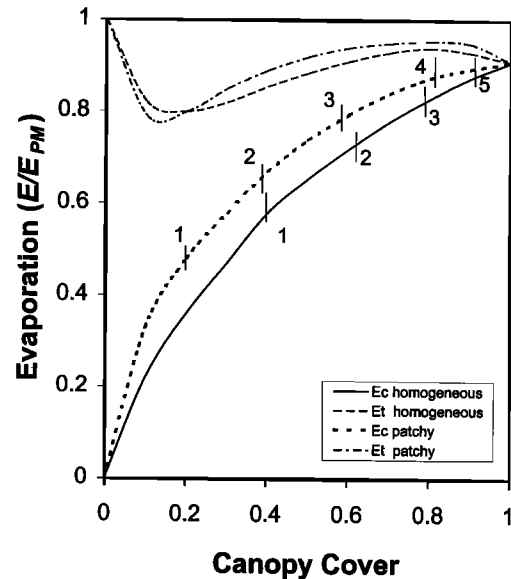
dry parts is less extreme. Extreme situations were preferred to determine the sensitivity of evaporation to the distribution of wetness. The third method to simulate partial canopy wetness assumes a horizontal alternation of wet and dry forest. Without advection the third method would be consistent with (3), so differences with (3) inform about the significance of small-scale advection on forest evaporation.

### 3. Simulation Results

#### 3.1. Evaporation From Closed, Wet Forest

The simulated vertical distribution of sensible and latent heat flux sources is shown in Figure 2. Most evaporation occurs near the top of the canopy. Absorbed radiation is a main source of energy for the evaporation of water. The remaining latent heat arises from the sink of sensible heat in the canopy. At the understory, evaporation is indistinguishable from the equilibrium value  $E = E_{eq} = 3.8 \text{ W m}^{-2}$ , matching to a sensible heat flux of  $1.5 \text{ W m}^{-2}$ . So the source of sensible heat flux changes sign near the ground surface. Owing to the change of sign, sensible heat is absorbed at a high mean level of 18.4 m as compared to the mean height of leaf area of 12.7 m. The mean source height for evaporation is 15.3 m, and the mean sink for momentum is located at 16.6 m height. The relatively high source level of the heat and momentum fluxes results mainly from the low aerodynamical resistance in the upper layers of the canopy. Differences in source height are not in line with the assumption of equal sources as made in the Penman-Monteith concept, showing that it is not self-evident to use  $E_{PM}$  for the calculation of rain-wetted forest.

Evaporation from a fully wet, closed forest is simulated as



**Figure 3.** Evaporation of wet forest versus canopy cover for homogeneous and patchy forest. Incomplete canopy cover of homogeneous forest arises from gaps between randomly spaced leaves. Numbers on the curves refer to the leaf area index. Shown are total evaporation ( $E_t$ ) and evaporation of the forest canopy ( $E_c$ ); the difference is understory evaporation.

$E = 0.90 E_{PM}$ . The result is influenced by the leaf boundary resistance or leaf width. Using  $l_w = 10^{-3} \text{ m}$ , a value representative for individual needles results in  $E = 1.36 E_{PM}$  and a further rise of the heat sources to the top of the canopy.

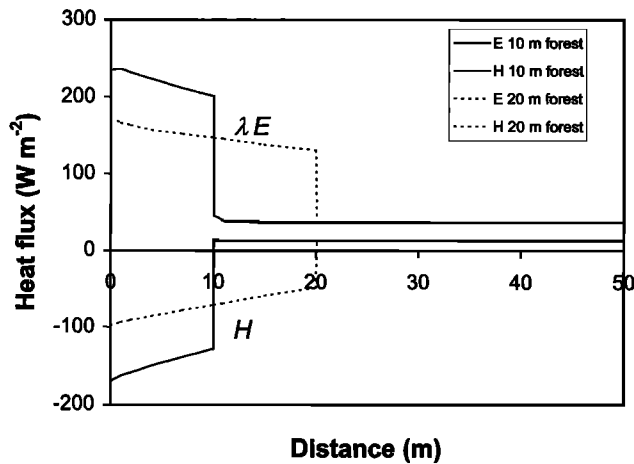
#### 3.2. Evaporation From Wet, Open Forest

Figure 3 shows simulated evaporation from forest with partial canopy cover. The total evaporation ( $E$ ) is separated into evaporation from the understory ( $E_s$ ) and the forest canopy ( $E_c$ ). Simulated understory evaporation for  $c = 0$  equals  $E_{PM}$  by definition, as evaporation from the understory surface is calculated with  $E_{PM}$ . Total evaporation  $E \approx E_{PM}$  for all values of canopy cover. This does not mean that  $E$  is independent of canopy cover, as the surface roughness and thus  $E_{PM}$  increases with canopy cover. By consequence,  $E$  increases from  $E = 60 \text{ W m}^{-2}$  for  $c = 0$  to  $E = 93 \text{ W m}^{-2}$  for  $c = 1$ . The increase of roughness and evaporation with forest density continues until  $\text{LAI} \sim 5$ . For  $\text{LAI} > 5$  the surface roughness and  $E_{PM}$  decrease slightly with further increasing leaf area as the wind penetrates less deeply into the forest, in agreement with simulations by Shaw and Pereira [1982]. Also, the ratio  $E/E_{PM}$  decreases slightly for  $\text{LAI} > 5$  owing to the decrease of wind velocity in the forest and increase of leaf boundary resistance. Evaporation from the forest canopy increases with canopy cover. The square root function

$$E_n = E_{c=1} c^{0.5} \quad (9)$$

falls in between the evaporation from the two forest representations, shown in Figure 3.

The horizontal variability of heat fluxes is shown in Figure 4 for patchy forest with  $c = 0.2$  and  $c = 0.4$ . With decreasing canopy cover, evaporation increases in the remaining forest. The highest evaporation is found at the wind-exposed edge (at distance = 0), and evaporation decreases slowly with distance

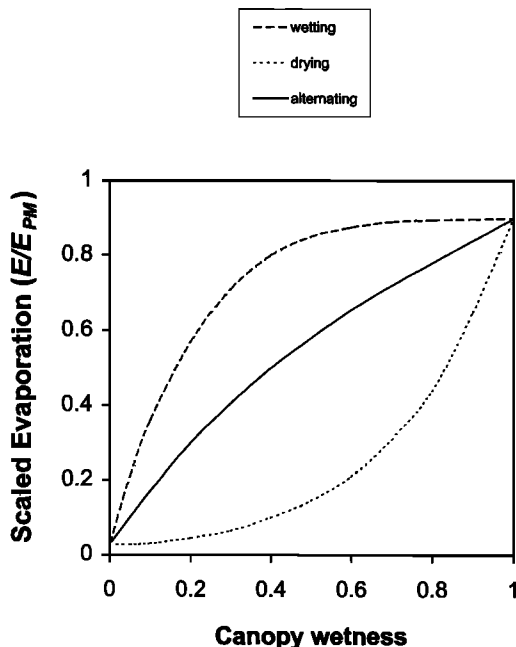


**Figure 4.** Horizontal source distribution of sensible and latent heat fluxes in patchy forest. An alternating pattern of 50 m is simulated, consisting of a forest strip of 10 m length (solid line) respectively 20 m length (dashed line), followed by grassland.

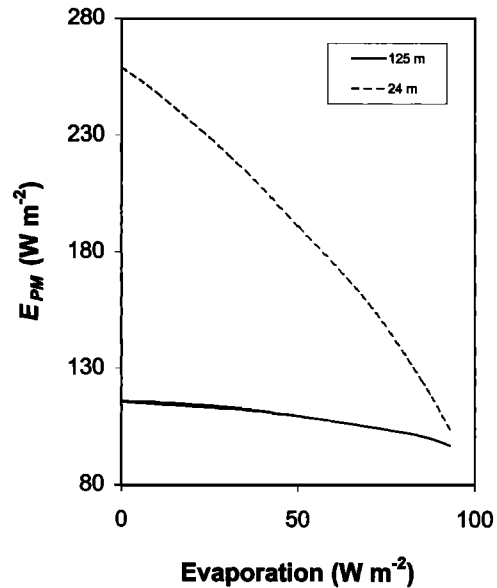
to the limit of  $93 \text{ W m}^{-2}$  that is simulated for homogeneous forest. Evaporation of the unforested patch is almost independent of canopy cover:  $\lambda E_s = 35 \text{ W m}^{-2}$  for  $c = 0.2$  and  $\lambda E_s = 34 \text{ W m}^{-2}$  for  $c = 0.4$ . These values are approximated by the equilibrium value  $\lambda E_{eq} = 34 \text{ W m}^{-2}$ . Near-equilibrium evaporation results from the low wind velocity between the tree lines: the wind velocity at 2 m height was simulated to be smaller than  $0.5 \text{ m s}^{-1}$ .

### 3.3. Evaporation From Partially Wet Forest

Scaled evaporation  $E/E_{PM}$  increases with wet fraction of the forest and is strongly sensitive to the distribution of wet and dry parts in the canopy (Figure 5). Scaled evaporation is mainly



**Figure 5.** Evaporation of partially wet forest using three distributions of wetness as described in sections 3.3 and 4.2.3.



**Figure 6.** The Penman-Monteith evaporation with zero surface resistance ( $E_{PM}$ ) calculated at two simulation heights as a function of actual evaporation. The variation in actual evaporation results from changes in the wet fraction of forest using the data of Figure 5 for horizontally alternating wetness.

dependent on the wetness of the upper leaves. By consequence, but not shown in Figure 5, the difference between the wetting and drying curves increases with LAI. Intermediate between the wetting and drying curves is the curve where partial wetness is simulated as a horizontal alternation of wet and dry forest parts. This curve is close to a straight line. As a straight line is expected in the absence of horizontal advection, the curve for alternating wetness implies that horizontal advection is of minor importance only. The small influence of horizontal advection does not mean that local fluxes are independent of advection (see Figure 4), but only that the horizontally averaged  $E/E_{PM}$  is not sensitive to horizontal heterogeneity in the forest canopy.

The curves in Figure 5 do not reach the origin, as a wet understorey is simulated. Understorey evaporation is increased by a factor of 2 to  $5.2 \text{ W m}^{-2}$  as compared to the completely wet situation, but the absolute difference with  $E_{eq}$  remains small.

The curves in Figure 5 are shown for the ratio between  $E$  and  $E_{PM}$ . This method of presentation might suggest that  $E_{PM}$  is independent of  $E$ . However, an increase of  $E$  increases the atmospheric humidity close to the forest and thus  $E_{PM}$  (Figure 6). Figure 6 was made with the input data of Figure 5, which are given at 125 m height. The curves show the resulting  $E_{PM}$  at two heights. The curves would intersect at  $E = E_{PM} = 97 \text{ W m}^{-2}$ . So only for  $E = E_{PM}$ ,  $E_{PM}$  is independent of height. When evaporation decreases,  $E_{PM}$  increases, and this increase is strongest near the evaporating surface. Figure 6 is used to validate the complementary relation (5). At 125 m height,  $E_{PM}$  is slightly sensitive to the evaporation of the forest:  $\varepsilon = -0.2$  as an increase of surface evaporation results in an increase of atmospheric stability and transport resistance. Note that at this height the atmospheric humidity deficit is fixed by the input data. Closer to the surface the atmospheric humidity deficit is increasingly affected by surface evaporation, resulting in a further enhancement of  $\varepsilon$ . At 24 m height the sensitivity  $\varepsilon = -1.8$

**Table 2.** Sensitivity of Evaporation to Model Parameters

| Input Data                          |                         |  |  | Output Data                |  |                            |                   |
|-------------------------------------|-------------------------|--|--|----------------------------|--|----------------------------|-------------------|
| Wind Velocity,<br>$\text{m s}^{-1}$ | Relative Humidity,<br>% | Incoming Radiation,<br>$\text{W m}^{-2}$ | Air Temperature,<br>$^{\circ}\text{C}$ | $A$ ,<br>$\text{W m}^{-2}$ | $E_{\text{PM}}$ ,<br>$\text{W m}^{-2}$ | $E$ ,<br>$\text{W m}^{-2}$ | $E/E_{\text{PM}}$ |
| 1                                   |                         |  |  | 71                         | 77                                     | 87                         | 1.13              |
| 2                                   |                         |  |  | 71                         | 85                                     | 87                         | 1.02              |
| 3                                   |                         |  |  | 72                         | 91                                     | 88                         | 0.97              |
| 5 <sup>a</sup>                      | 90 <sup>a</sup>         | 100 <sup>a</sup>                         | 20 <sup>a</sup>                        | 72 <sup>a</sup>            | 104 <sup>a</sup>                       | 93 <sup>a</sup>            | 0.90              |
| 7                                   |                         |  |  | 72                         | 119                                    | 100                        | 0.84              |
| 10                                  |                         |  |  | 73                         | 152                                    | 116                        | 0.76              |
| 20                                  |                         |  |  | 74                         | 289                                    | 172                        | 0.60              |
|                                     | 85                      |  |  | 74                         | 114                                    | 103                        | 0.90              |
|                                     | 95                      |  |  | 69                         | 81                                     | 74                         | 0.91              |
|                                     | 99                      |  |  | 67                         | 54                                     | 53                         | 1.02              |
|                                     |                         | 0  |  | UC <sup>b</sup>            | UC <sup>b</sup>                        | UC <sup>b</sup>            | UC <sup>b</sup>   |
|                                     |                         | 10                                       |  | -36                        | 51                                     | 13                         | 0.25              |
|                                     |                         | 50                                       |  | 30                         | 52                                     | 47                         | 0.90              |
|                                     |                         | 200                                      |  | 158                        | 187                                    | 170                        | 0.91              |
|                                     |                         | 300                                      |  | 246                        | 261                                    | 240                        | 0.92              |
| 10                                  |                         | 0  |  | -48                        | 76                                     | 46                         | 0.61              |
| 10                                  |                         | 200                                      |  | 159                        | 203                                    | 186                        | 0.92              |
| 10                                  | 95                      | 200                                      |  | 156                        | 173                                    | 150                        | 0.87              |
|                                     |                         |  | 10                                     | 122                        | 129                                    | 115                        | 0.89              |

<sup>a</sup> Standard simulation, also used when no input data were given.

<sup>b</sup>UC, unstable calculation; Simulation resulted in such a strong atmospheric stability that the wind velocity in the forest almost vanished.

even exceeds the sensitivity resulting from the complementary relation ( $\varepsilon = -1$ ).

### 3.4. Environmental Control of Wet Forest Evaporation

As evaporation from wet surfaces depends on available energy ( $A$ ) and transport  $(e_a^* - e_a)/r_a$ , see (1), incoming radiation (input for  $A$ ), relative humidity (input for  $(e_a^* - e_a)$ ), and wind velocity (input for  $r_a$ ) may influence evaporation. The simulation uses input data at the upper layer of the model at 125 m height, enabling the air near the forest to adjust to the surface conditions. Input data and results are given in Table 2.

The simulations show a strong control of available energy on evaporation. Comparing the simulations with  $R_s = 50$  and  $R_s = 200 \text{ W m}^{-2}$  results in  $E = 0.961 A + 23.6 \text{ (W m}^{-2}\text{)}$ , so  $dE/dA = 0.961$ . Other combinations of  $R_s$  result in similar relations. Evaporation for  $A = 0$  is maintained by atmospheric transport of sensible heat. The simulated sensitivity  $dE/dA = 0.961$  exceeds the sensitivity of equilibrium evaporation,  $dE_{eq}/dA = 0.69$ . The formula  $dE/dA \neq dE_{eq}/dA$  means that atmospheric transport is affected by available energy and in this case that atmospheric transport is enhanced with increasing available energy.

The sensitivity of evaporation to other environmental parameters is limited: An increase of wind velocity by a factor of 20 (from 1 to 20  $\text{m s}^{-1}$ ) results in a doubling of evaporation (from 87 to 172  $\text{W m}^{-2}$ ) and an increase of a factor of 15 in atmospheric humidity deficit (relative humidity from 99 to 85%) also results in an enhancement of evaporation by a factor of 2 (from 53 to 103  $\text{W m}^{-2}$ ).

## 4. Discussion

### 4.1. Model Assumptions

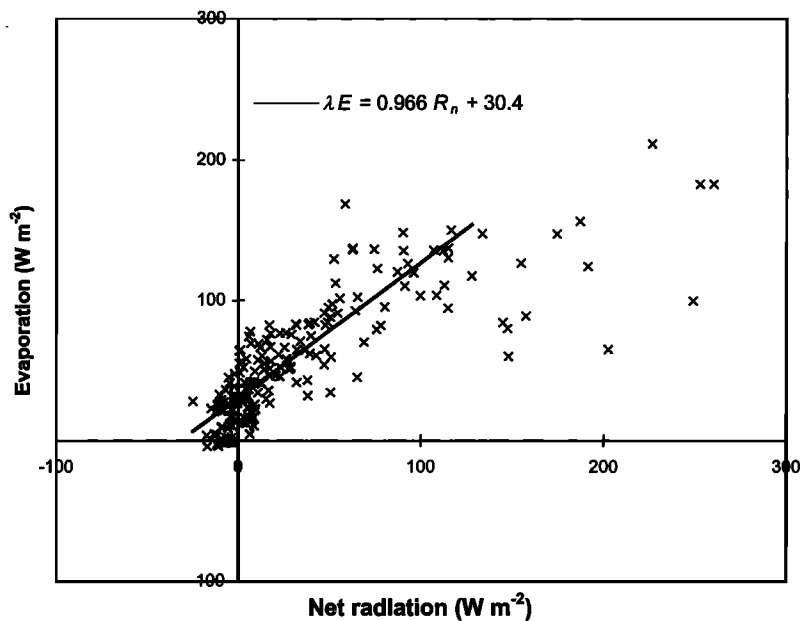
Major assumptions in the simulation of evaporation from rain-wetted forest with the model of *Klaassen* [1992] are adjustment and stationarity. Adjustment means that the atmo-

spheric profiles of wind velocity, temperature, and humidity are iterated until variations after successive iterations are negligible. For a homogeneous surface, adjustment results in horizontally and vertically constant fluxes in the atmospheric surface layer. For a heterogeneous surface, adjustment has been defined as adjustment to the horizontally averaged fluxes of the surface. Adjustment does not occur just after a single transition between low vegetation and forest. This situation was evaluated by *Klaassen et al.* [1996]. The main conclusion was that the evaporation rate is increased and the water storage capacity of the canopy decreased near wind exposed forest edges.

Stationarity means that time derivatives are neglected, such as heat release by cooling ( $Q = C_H \partial T / \partial t$ ), which would add to the available energy for evaporation. A rough upper estimate is given of the magnitude of heating effects. It seems reasonable to assume that the canopy will cool during the initial phase of a rain shower. *Moore and Fisch* [1986] determined a heat coefficient  $C_H = 33 \text{ W m}^{-2} ^{\circ}\text{C}^{-1} \text{ h}^{-1}$  in the Amazonian rain forest. Taking this value and assuming a temperature decrease of  $3^{\circ}\text{C}$  during the first hour of rain results in a heat source  $Q = 99 \text{ W m}^{-2}$ . It might also be speculated that the temperature of rainwater would increase when it comes in contact with the warmer surface. Assuming a rain intensity of  $10 \text{ mm h}^{-1}$  and a warming of the rainwater by  $3^{\circ}\text{C}$  would result in a sink of  $35 \text{ W m}^{-2}$ . Although these estimates are very rough, they show that processes of heat release might influence the available energy and thus the evaporation from rain-wetted forest. As the effects are most pronounced during the initial phase of precipitation, the wetting-up phase will not be analyzed.

### 4.2. Evaluation of Simulated Results

**4.2.1. Sensitivity of evaporation to net radiation.** The simulated sensitivity to available energy is compared to measurements above a dense Douglas fir forest at Speult ( $52^{\circ}15'\text{N}$ ,  $5^{\circ}41'\text{W}$ , 52 m above sea level) in the Netherlands. *Bouten et al.*



**Figure 7.** Measured half-hourly evaporation versus net radiation for Douglas fir forest when water storage in the canopy exceeds 2.0 mm on a maximum value of 2.5 mm. The line is a regression through data for  $R_n < 130 \text{ W m}^{-2}$ .

[1991, 1996], *Klaassen et al.* [1998], and *Bosveld and Bouten* [1999] have described the measurements. All half-hour-averaged data are used when water storage on the canopy exceeds 2.0 mm. As the maximum water storage capacity was measured by microwave transmission as 2.5 mm, Figure 7 shows data when the canopy was (almost) saturated. Evaporation was calculated from vertical gradients and from eddy correlation measurements. The averaged value was used when both methods yielded realistic results. Evaporation is related to net radiation, as this parameter is directly measured and thus more accurately determined than available energy. The measurements for  $R_n < 130 \text{ W m}^{-2}$  are fitted with  $E = 0.966R_n + 30.4 \text{ (W m}^{-2}\text{)}$ , with  $R^2 = 0.72$  showing that net radiation is the main factor controlling evaporation from wet forest. The measurements are in excellent agreement with the simulated result ( $E = 0.961A + 23.6 \text{ W m}^{-2}$  and  $A = 0.992R_n$ ). The data for  $R_n < 130 \text{ W m}^{-2}$  are representative for rain. The experimental data for  $R_n > 130 \text{ W m}^{-2}$  show more scatter and less sensitivity to net radiation. These high-radiation values are representative for dry periods after rain when drying has just started and the ratio  $E/E_{PM}$  quickly decreases [*Bosveld and Bouten*, 1999].

A strong control of available energy on evaporation from rain-wetted forest seems contradictory to the analysis made by *Jarvis and McNaughton* [1986]. They show that wet forest is well coupled to the lower atmosphere, resulting in a strong control of wind and humidity on evaporation. The deviating result of the present simulation is caused by the high level of 125 m height where data were used as input. The agreement with measurements justifies the use of input data at a level well above most experimental observations. Forest evaporation is strongly coupled to the wind and humidity deficit close to the forest, but the coupling reduces quickly with increasing height owing to the slightly stable atmospheric stratification.

Measurements and simulations both show the evaporation to increase even stronger with available energy than calculated

using equilibrium evaporation. According to the simulations this sensitivity of evaporation to available energy is caused by a decrease of atmospheric stability and atmospheric resistance with increasing available energy. The poor sensitivity of evaporation to wind and humidity at given available energy is caused by negative feedback between evaporation and near-surface humidity deficit.

**4.2.2. The Penman-Monteith estimate of wet forest evaporation.** The simulations show small differences with the Penman-Monteith estimate of evaporation. The sensitivity of evaporation to leaf width is limited in real forests, as needles are jointed to a twig with almost still air in between the needles. Another deviation from  $E_{PM}$  is simulated for heavy storms (wind velocity  $20 \text{ m s}^{-1}$ ). When leaves are exposed to high wind velocities, *Brenner and Jarvis* [1995] found that (6) overestimates the leaf layer boundary resistance. A lower leaf boundary resistance during storms would result in a higher evaporation and thus to a smaller difference between  $E$  and  $E_{PM}$ . So the validity of  $E_{PM}$  for the evaporation from wet, closed forest is expected to be even better than suggested by Table 2 and agrees with other multilayer model simulations [*Watanabe and Mizutani*, 1996; *Bouten et al.*, 1996].

The theoretical result that  $E_{PM}$  is useful to estimate evaporation from wet forest is compared to published observations. The Speult data are not useful to validate this thesis as splashing raindrops frequently wetted the humidity sensor; so  $E_{PM}$  could not be calculated with sufficient reliability. The measurements by *Stewart* [1977] were selected on consistency with  $E_{PM}$ . As a consequence, these data are also not useful for validation. The measurements of *Lindroth* [1993] and *Mizutani et al.* [1997] indicate that  $E_{PM}$  is a useful approximation when the canopy is fully wet. The measurements by *Gash et al.* [1999] show  $E < E_{PM}$ . Their explanation that evaporation would be reduced owing to incomplete canopy cover is not satisfactory, as meteorological measurements inform us about total evaporation. Figure 4 shows that total evaporation ( $E/E_{PM}$ ) is not



sensitive to canopy cover. Instead, it seems more reasonable to assume that the canopy was not fully wet all the time, as canopy wetness was simulated with a model that neglects drainage before saturation of the canopy. Measured evaporation by *Lankreijer et al.* [1999] was also below simulated  $E_{PM}$ , and again the result is explained by neglect of drainage and overestimation of canopy wetness in their simulations. It is concluded that  $E_{PM}$  is useful to calculate the evaporation of wet forest, but a careful check is required whether the forest is fully wet.

**4.2.3. Sensitivity of evaporation to forest wetness.** The difference between the three curves in Figure 5 suggests that a single relation between evaporation and wetness is questionable. Instead, hysteresis characterizes the different evaporation rates at the onset of wetting and drying. For instance, evaporation is expected to reduce quickly during the initial phase of drying, as the upper leaves dry most quickly owing to their relatively good coupling to the atmospheric conditions just outside the forest. The same process results in a quick drying of the best coupled parts of the leaves. So quick drying of well-coupled canopy parts acts on different scales and should always result in a more than proportional decrease of evaporation with wetness during the onset of drying.

A more than proportional reduction of evaporation with storage at the onset of drying agrees with measurements of *Calder and Wright* [1986] and *Bosveld and Bouten* [1999] but disagrees with measurements of *Teklehaimanot and Jarvis* [1991]. In the latter study, however,  $E_{PM}$  was calculated assuming negligible available energy. By contrast, Figure 7 shows that available energy is the most important factor determining evaporation from wet forest. As available energy generally increases during daytime after the rain event, it is speculated that the data of *Teklehaimanot and Jarvis* [1991] agree with previous measurements that evaporation reduces more than proportional to storage when rain has stopped. On the scale of leaves, *Larsson* [1981] showed evaporation to decrease proportional to storage for one *Salix* species and a more than proportional decrease for another *Salix* species. So experimental evidence indicates that evaporation reduces stronger than proportional to  $C/S$  just after rain. The reduction of evaporation might depend on forest species, as the distribution of water on the leaves varies [e.g., *Horton*, 1919] and may influence the evaporation rate.

The suggestion that evaporation reduces less than proportional to storage during wetting up has not been validated because of the uncertainties in available energy, as discussed in section 4.1. As the simulations show that  $E/E_{PM}$  can be smaller as well as larger than  $C/S$  and that the observations are also ambiguous, the Rutter assumption  $E/E_{PM} = C/S$  is still recommended as a provisionally acceptable overall relation.

**4.2.4. Sensitivity of evaporation to canopy cover.** The simulated square root increase of evaporation with canopy cover (equation (9)) means that (2), assuming canopy evaporation to increase proportional with canopy cover, would clearly underestimate evaporation. The small difference between the two forest representations shows that within-canopy advection is of minor influence on the spatially averaged evaporation. The similarity in evaporation resulting from two strongly deviating representations of open forest indicates that the simulated result, a square root increase of evaporation with canopy cover, can also be transferred to real, open forest with a more complex architecture.

Direct measurements are not available to test the square

root relation. However, indirect evidence for the evaporation to exceed proportionality with canopy cover is found in thinning experiments by *Veracón and Lopez* [1976], *Aussenac et al.* [1982], *Crockford and Richardson* [1990], *Teklehaimanot et al.* [1991], and *Baümler and Zech* [1997]. All these experiments show that rainfall interception decreases less than proportional with the thinned area, a result that can now be explained by the increase of evaporation from the remaining forested area. The simulated result that evaporation from the wet understorey can be calculated with the equilibrium evaporation is physically realistic as long as the understorey is sheltered from the wind.

#### 4.3. Simulation of Evaporation in Rainfall Interception Models

The model of *Gash* [1979] is taken as a reference, as this model is widely used to simulate rainfall interception. With this model, rainfall interception by dense forest is simulated well, but the interception by open forest is overestimated [*Lankreijer et al.*, 1993, 1999; *Gash et al.*, 1995, 1999]. As storage of water is restricted in open forest, the overestimation of interception of open forest is explained by an overestimation of the evaporation rate in the reference model. Two reasons have been suggested to explain the overestimation. The explanation by *Lankreijer et al.* [1999] that the overestimation is caused by an internal resistance to water vapor transport does not agree with the present theoretical analysis. Also, a supplementary resistance cannot explain the success of the reference model to simulate rainfall by dense forest. The suggestion by *Gash et al.* [1999] that evaporation from wet forest is proportional to canopy cover does not agree with the present analysis as canopy evaporation exceeds this value. Moreover, his meteorological measurements are representative for the total evaporation, and total evaporation is shown to be hardly dependent on canopy cover. So an overestimation of evaporation from open forest using the reference model is not caused by an additional internal transport resistance or by the suggested correction for incomplete canopy cover.

An alternative method to restrict the evaporation rate in interception models is to enhance the correction for incomplete canopy wetness. The present analysis still suggests to use the common correction ( $E/E_{PM} = C/S$ , equation (6)). A larger correction also results from a decrease in canopy wetness ( $C/S$ ), which can result from a decrease in  $C$  or an increase in  $S$ . The maximum storage capacity  $S$  is commonly determined indirectly from throughfall measurement and results in an underestimation of  $S$  when drainage from the canopy occurs before the canopy is saturated [*Klaassen et al.*, 1998]. Three processes cause drainage before saturation: raindrops splashing on wetted parts [*Calder*, 1986], slow saturation of barks and undersides of leaves [*Herwitz*, 1985], and upper leaves sheltering lower leaves. So drainage results a lower canopy wetness and a lower evaporation rate. The low evaporation rate in combination with the limited water storage capacity explains the measured low-rainfall interception of open forest. For dense forest, maximum storage is larger and results in an increase of interception. The realistic simulation of interception of dense forest using the reference model is thus explained by compensation errors, resulting from an overestimation of the evaporation rate and an underestimation of storage [*Klaassen et al.*, 1998].

The present analysis argues that rainfall interception models improve from a simulation of drainage during incomplete canopy wetness. Several methods already exist to simulate drain-

age: Rutter *et al.* [1971] and Massman [1980] assume the drainage rate to increase with canopy wetness. By setting drainage also proportional to the rainfall rate the wetting efficiency of rainfall, defined as the increase of canopy storage per unit of rainfall, is made dependent on canopy storage [Calder, 1986; Liu, 1996]. Also, a larger water storage capacity can easily be included in rainfall interception models. However, it would become more difficult to determine such a water storage capacity from the common measurements of the difference in rainfall outside and inside the forest canopy [Klaassen *et al.*, 1998]. Instead, a supplementary measurement is needed to distinguish between the components storage and evaporation during rain. Recent experiments have shown the feasibility to measure the parameters storage [e.g., Calder and Wright, 1986; Dunin *et al.*, 1988; Teklehaimanot and Jarvis, 1991; Bouten *et al.*, 1991] and evaporation during rain [e.g., Stewart, 1977; Lindroth, 1993; Mizutani *et al.*, 1997; Gash *et al.*, 1999; Lankreijer *et al.*, 1999] during field experiments. The input data for a more realistic simulation of rainfall interception can now be obtained.

## 5. Concluding Remarks

Using a multilayer canopy representation, it was shown theoretically that it is not self-evident to calculate evaporation from wet forest with  $E_{PM}$ , the representative value for a homogeneous single layer canopy. A higher evaporation can occur when the heat sources are located near the top of the canopy. On the other hand, the leaf boundary resistance results in a lower evaporation than calculated with  $E_{PM}$ . As these opposing effects almost compensate, it is concluded that  $E_{PM}$  is a useful empirical approximation for most studies on rainfall interception. Corrections on  $E_{PM}$  are needed for partial canopy cover and wetness. As wetness is defined as the water covered fraction of the canopy, it should be noted partial wetness might even occur when the canopy is saturated in case of hydrophobic leaves.

Evaporation of partially wet canopies is very sensitive to the distribution of wetness. Yet for most models on rainfall interception a single relation between evaporation and wetness is to be preferred. The common proportionality between  $E/E_{PM}$  and  $C/S$  seems useful for most hydrological applications.

During incomplete wetness,  $E_{PM}$  increases as vapor pressure deficit and atmospheric transport resistance increase with reduced evaporation. So  $E_{PM}$  is not a constant but rather is a state variable, which appears to depend on measurement height. As  $E_{PM}$  is very sensitive to atmospheric vapor pressure deficit and this parameter is difficult to determine accurately during rain, evaporation might alternatively be calculated from the net radiation or available energy. It is recommended to test the observed relation between evaporation and net radiation in other climates.

Studies on rainfall interception are commonly based on the difference in rainfall outside and inside the canopy. Such measurements should be used in combination with an estimate of evaporation from the canopy only. By contrast, meteorological measurements above the canopy relate to total evaporation from canopy and understorey. Therefore it is important to distinguish between these components of evaporation, in particular for canopies with incomplete canopy cover where the difference between total and canopy evaporation is large. Evaporation from wet understorey can be approximated by the local equilibrium evaporation when canopy cover is larger than

20%. Evaporation from wet canopies appears proportional to the square root of canopy cover.

The simulation of interception improves from a parameterization of evaporation with values below  $E_{PM}$ . The present analysis shows that the reduction of evaporation is most likely caused by a smaller than expected wetness of the canopy during rainfall. A listing has been given of useful models to simulate a smaller canopy wetness as well as measurement methods to determine the necessary input data.

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## References

- Aussenac, G., A. Granier, and R. Naud, Influence of thinning on growth and water balance, *Can. J. For. Res.*, 12(2), 222–231, 1982.
- Bäumler, R., and W. Zech, Atmospheric deposition and impact of forest thinning on the throughfall of mountain forest ecosystems in the Bavarian Alps, *For. Ecol. Manage.*, 95, 243–251, 1997.
- Bosveld, F. C., and W. Bouten, Evaporation and transpiration reduction of a partially wet Douglas-fir forest, *KNMI Preprints 99-19*, 29 pp., K. Ned. Meteorol. Inst., De Bilt, Netherlands, 1999.
- Bouchet, R. J., Evapotranspiration réelle et potentielle, signification climatique, *IASH Publ.*, 62, 134–142, 1963.
- Bouten, W., P. J. F. Swart, and E. de Water, Microwave transmission, a new tool in forest hydrological research, *J. Hydrol.*, 124, 119–130, 1991.
- Bouten, W., M. G. Schaap, J. Aerts, and A. W. M. Vermetten, Monitoring and modelling canopy water storage amounts in support of atmospheric deposition studies, *J. Hydrol.*, 181, 305–321, 1996.
- Brenner, A. J., and P. G. Jarvis, A heated leaf replica technique for determination of leaf boundary layer conductance in the field, *Agric. For. Meteorol.*, 72, 261–275, 1995.
- Calder, I. R., A stochastic model of rainfall interception, *J. Hydrol.*, 89, 65–71, 1986.
- Calder, I. R., and I. R. Wright, Gamma ray attenuation studies of interception from Sitka spruce: Some evidence for an additional transport mechanism, *Water Resour. Res.*, 22, 409–417, 1986.
- Choudhury, B. J., and N. E. DiGirolamo, A biophysical process-based estimate of global land surface evaporation using satellite and ancillary data, I, Model description and comparison with observations, *J. Hydrol.*, 205, 164–185, 1998.
- Crockford, R. H., and D. P. Richardson, Partitioning of rainfall in an eucalypt forest and pine plantation in southern Australia, IV, The relationship of interception and storage capacity, the interception of these forests, and the effect on interception of thinning the pine plantation, *Hydrol. Processes*, 4, 168–188, 1980.
- Dunin, F. X., E. M. O'Loughlin, and W. Reyenga, Interception loss from eucalypt forest: Lysimeter determination of hourly rates for long term evaluation, *Hydrol. Processes*, 2, 315–329, 1988.
- Gash, J. H. C., An analytical model of rainfall interception by forests, *Q. J. R. Meteorol. Soc.*, 105, 43–55, 1979.
- Gash, J. H. C., C. R. Lloyd, and G. Lachaud, Estimating sparse forest rainfall interception with an analytical model, *J. Hydrol.*, 170, 79–86, 1995.
- Gash, J. H. C., F. Valente, and J. S. Davis, Estimates and measurements of evaporation from wet, sparse pine forest in Portugal, *Agric. For. Meteorol.*, 94, 149–158, 1999.
- Herwitz, S. R., Interception storage capacities of tropical rainforest trees, *J. Hydrol.*, 77, 237–252, 1985.
- Horton, R. E., Rainfall interception, *Mon. Weather Rev.*, 47, 603–623, 1919.
- Hutjes, R. W. A., A. Wierda, and A. W. L. Veen, Rainfall interception in the Tai forest, Ivory Coast: Application of two simulation models to a humid tropical system, *J. Hydrol.*, 114, 259–275, 1990.
- Jarvis, P. G., and J. W. Leverenz, Productivity of temperature, deciduous and evergreen forests, in *Encyclopedia of Plant Physiology*, vol.

- 12D, *Physiological Plant Ecology Responses to the Physical Environment*, edited by O. L. Lange, P. S. Nobel, C. B. Osmond, and H. Ziegler, pp. 233–280, Springer-Verlag, New York, 1983.
- Jarvis, P. G., and K. G. McNaughton, Stomatal control of transpiration: Scaling up from leaf to region, *Adv. Ecol. Res.*, 15, 1–49, 1986.
- Klaassen, W., Average fluxes from heterogeneous vegetated regions, *Boundary Layer Meteorol.*, 58, 329–354, 1992.
- Klaassen, W., H. J. M. Lankreijer, and A. W. L. Veen, Rainfall interception near a forest edge, *J. Hydrol.*, 185, 349–361, 1996.
- Klaassen, W., F. Bosveld, and E. de Water, Water storage and evaporation as constituents of rainfall interception, *J. Hydrol.*, 212–213, 36–50, 1998.
- Lankreijer, H. J. M., M. J. Hendriks, and W. Klaassen, A comparison of models simulating rainfall interception of forests, *Agric. For. Meteorol.*, 64, 187–199, 1993.
- Lankreijer, H. J. M., A. Lundberg, A. Grelle, A. Lindroth, and J. Seibert, Evaporation and storage of intercepted rain analysed by comparing two models applied to a boreal forest, *Agric. For. Meteorol.*, 98–99, 595–604, 1999.
- Larsson, S., Influence of intercepted water on transpiration and evaporation of *Salix*, *Agric. Meteorol.*, 23, 331–338, 1981.
- Li, Z., D. R. Miller, and J. D. Lin, A first order closure scheme to describe countergradient momentum transport in plant canopies, *Boundary Layer Meteorol.*, 25, 25–41, 1985.
- Lindroth, A., Aerodynamic and canopy resistance of short-rotation forest in relation to leaf area index and climate, *Boundary Layer Meteorol.*, 66, 265–279, 1993.
- Liu, S., A new model for the prediction of rainfall interception in forest canopies, *Ecol. Modell.*, 99, 151–159, 1996.
- Loustau, D., P. Berbigier, and A. Granier, Interception loss, throughfall and stemflow in a maritime pine stand, II, An application of Gash's analytical model of interception, *J. Hydrol.*, 138, 469–485, 1992.
- Massman, W. J., Water storage on forest foliage: A general model, *Water Resour. Res.*, 16, 210–216, 1980.
- Mizutani, K., K. Yamanoi, T. Ikeda, and T. Watanabe, Applicability of the eddy correlation method to measure sensible heat transfer to forest under rainfall conditions, *Agric. For. Meteorol.*, 86, 193–203, 1997.
- Moore, C. J., and G. Fisch, Estimating heat storage in Amazonian tropical forest, *Agric. For. Meteorol.*, 38, 147–169, 1986.
- Pearman, G. I., H. L. Weaver, and C. Tanner, Boundary layer heat transfer coefficients, *Agric. Meteorol.*, 10, 83–92, 1972.
- Rutter, A. J., K. A. Keshaw, P. C. Robins, and A. J. Morton, A predictive model of rainfall interception in forests, I, Derivation of the model from observations in a plantation of corsican pine, *Agric. Meteorol.*, 9, 367–384, 1971.
- Shaw, R. H., and A. R. Perreira, Aerodynamical roughness of a plant canopy: A numerical experiment, *Agric. Meteorol.*, 26, 51–65, 1982.
- Stewart, J. B., Evaporation from the wet canopy of a pine forest, *Water Resour. Res.*, 13, 915–921, 1977.
- Teklehaimanot, Z., and P. G. Jarvis, Direct measurement of evaporation of intercepted water from forest canopies, *J. Appl. Ecol.*, 28, 603–618, 1991.
- Teklehaimanot, Z., P. G. Jarvis, and D. C. Ledger, Rainfall interception and boundary layer conductance in relation to tree spacing, *J. Hydrol.*, 123, 261–278, 1991.
- Valente, F., J. S. David, and J. H. C. Gash, Modelling interception loss for two sparse eucalypt and pine forests in central Portugal using reformulated Rutter and Gash analytical models, *J. Hydrol.*, 190, 141–162, 1997.
- Veracion, V. P., and A. C. B. López, Rainfall interception in a thinned Benguet pine forest, *Sylvatrop.*, 1(2), 128–134, 1976.
- Watanabe, T., and K. Mizutani, Model study on micrometeorological aspects of rainfall interception over an evergreen broad-leaved forest, *Agric. For. Meteorol.*, 80, 195–214, 1996.

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